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Energy partitioning between latent and sensible heat flux during the warm season at FLUXNET sites

Kell B. Wilson,¹ Dennis D. Baldocchi,² Marc Aubinet,³ Paul Berbigier,⁴ Christian Bernhofer,⁵ Han Dolman,⁶ Eva Falge,⁷ Chris Field,⁸ Allen Goldstein,² Andre Granier,⁹ Achim Grelle,¹⁰ Thorgeirsson Halldor,¹¹ Dave Hollinger,¹² Gabriel Katul,¹³ B. E. Law,¹⁴ Anders Lindroth,¹⁵ Tilden Meyers,¹ John Moncrieff,¹⁶ Russ Monson,¹⁷ Walter Oechel,¹⁸ John Tenhunen,⁷ Riccardo Valentini,¹⁹ Shashi Verma,²⁰ Timo Vesala,²¹ and Steve Wofsy²²

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[1] The warm season (mid-June through late August) partitioning between sensible (H) and latent (LE) heat flux, or the Bowen ratio ($\beta = H/LE$), was investigated at 27 sites over 66 site years within the international network of eddy covariance sites (FLUXNET). Variability in β across ecosystems and climates was analyzed by quantifying general climatic and surface characteristics that control flux partitioning. The climatic control on β was quantified using the climatological resistance (R_i), which is proportional to the ratio of vapor pressure deficit (difference between saturation vapor pressure and atmospheric vapor pressure) to net radiation (large values of R_i decrease β). The control of flux partitioning by the vegetation and underlying surface was quantified by computing the surface resistance to water vapor transport (R_c , with large values tending to increase β). There was a considerable range in flux partitioning characteristics (R_c , R_i and β) among sites, but it was possible to define some general differences between vegetation types and climates. Deciduous forest sites and the agricultural site had the lowest values of R_c and β (0.25–0.50). Coniferous forests typically had a larger R_c and higher β (typically between 0.50 and 1.00 but also much larger). However, there was notable variability in R_c and R_i between coniferous site years, most notably differences between oceanic and continental climates and sites with a distinct dry summer season (Mediterranean climate). Sites with Mediterranean climates generally had the highest net radiation, R_c , R_i , and β . There was considerable variability in β between grassland site years, primarily the result of interannual differences in soil water content and R_c . **INDEX TERMS:** 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 1818 Hydrology: Evapotranspiration; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; 1851 Hydrology: Plant ecology; 3307 Meteorology and Atmospheric Dynamics: Boundary layer processes; **KEYWORDS:** eddy covariance, evaporation, ecohydrology, hydrology

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¹Atmospheric Turbulence and Diffusion Division, NOAA, Oak Ridge, Tennessee, USA.

²Department of Environmental Science Policy and Management, University of California at Berkeley, Berkeley, California, USA.

³Unité de Physique, Faculté des Sciences Agronomiques de Gembloux, Gembloux, Belgium.

⁴Unité de Bioclimatologie, INRA Bourdeaux, Gazinet, France.

⁵Institute für Hydrologie und Meteorologie, Tharandt, Germany.

⁶Alterra, Wageningen, Netherlands.

⁷Pflanzenökologie, University Bayreuth, Bayreuth, Germany.

⁸Department of Plant Biology, Carnegie Institution of Washington, Stanford, California, USA.

⁹Unité Ecophysiologie Forestières, Institut Nationale de la Recherche Agronomique, Champenoux, France.

¹⁰Department of Ecology and Environmental Research, Swedish University of Agricultural Sciences, Uppsala, Sweden.

¹¹Agricultural Research Institute, Department of Environmental Research, Reykjavik, Iceland.

¹²USDA Forest Service, Northeastern Forest Experiment Station, Durham, New Hampshire, USA.

¹³School of the Environment, Duke University, Durham, North Carolina, USA.

¹⁴College of Forestry, Oregon State University, Corvallis, Oregon, USA.

¹⁵Department of Ecology and Environmental Research, Swedish University of Agricultural Sciences, Uppsala, Sweden.

¹⁶Department of Ecology and Environmental Research, University of Edinburgh, Edinburgh, UK.

¹⁷Department of Environmental, Population, and Organismic Biology, University of Colorado, Boulder, Colorado, USA.

¹⁸Global Change Research Group, San Diego State University, San Diego, California, USA.

¹⁹Department of Forest Science and Resources, University of Tuscia, Viterbo, Italy.

²⁰School of Natural Resource Sciences, University of Nebraska, Lincoln, Lincoln, Nebraska, USA.

²¹Department of Physical Sciences, University of Helsinki, Helsinki, Finland.

²²Department of Earth and Planetary Sciences, Harvard University, Cambridge, Massachusetts, USA.

1. Introduction

[2] The partitioning between ecosystem latent (LE) and sensible (H) heat fluxes is critical in determining the hydrological cycle, boundary layer development, weather and climate. Two important issues concerning flux partitioning are the variability of partitioning across different climates and ecosystems and the mechanisms for this variability. Energy partitioning at the surface is a complex function of longer-term interactions between biogeochemical cycling, disturbance, and climate, and shorter-term interactions between plant physiology and the development of the atmospheric boundary layer. For example, on longer time scales, leaf nitrogen, photosynthetic capacity, leaf area, canopy structure and maximum stomatal conductance of individual species have adapted to the local climate demand and the supply of nutrients [e.g., *Field and Mooney*, 1986; *Woodward*, 1987; *Neilson*, 1995; *Baldocchi and Meyers*, 1998]. On shorter time scales, stomatal conductance responds to the local environment, which changes as the atmospheric boundary layer develops [*Jacobs and de Bruin*, 1992; *Raupach*, 1998; *Wilson et al.*, 1999].

[3] The “big leaf” or “single-source” idealization of a canopy [*Monteith*, 1965; *Raupach and Finnigan*, 1988], which depicts the surface characteristics using a limited parameter set, is a method for simplifying the effects these complex processes have in controlling surface flux characteristics. Because of three-dimensional and nonlinear canopy processes, single-source models are not always an acceptable modeling approach [*De Pury and Farquhar*, 1997]. Nevertheless, the “big leaf” idealization is a practical approach for quantifying the atmospheric and vegetation controls of ecosystem energy fluxes without a detailed knowledge of complexity at each site. The “big leaf” paradigm parameterizes the canopy as a one-dimensional source for the turbulent exchange of fluxes [*Monteith*, 1965]:

$$H = \frac{\rho C_p (T_s - T_a)}{R_a} \quad (1)$$

$$LE = \frac{\rho C_p \delta_e / \gamma}{R_a + R_c} \quad (2)$$

where ρ is the density of air (kg m^{-3}); C_p is the specific heat of air ($1005 \text{ J kg}^{-1} \text{ K}^{-1}$); T_s and T_a are the “surface” and air temperatures ($^{\circ}\text{C}$); R_a is the atmospheric resistance to water vapor transport (s m^{-1}); δ_e is the atmospheric vapor pressure deficit (Pa); γ is the psychrometer constant ($\approx 64 \text{ Pa K}^{-1}$); R_c is the surface resistance to water vapor transport (s m^{-1}). Using the Penman-Monteith approximation [*Monteith*, 1965] to equations (1) and (2), *Jarvis et al.* [1976] expressed the ratio between H and LE (the Bowen ratio, β) as:

$$\beta = \frac{1 + (R_c/R_a) - (R_i/R_a)}{(s/\gamma) + (R_i/R_a)} \quad (3)$$

where the climatological resistance (R_i) is:

$$R_i = \frac{\rho C_p \delta_e}{\gamma A} \quad (4)$$

A is the available energy ($R_n - G$), and R_n is the net radiation, and G is the change in heat storage in the soil and canopy. s is the slope of the saturation vapor pressure curve with respect to temperature at a specified temperature (Pa K^{-1}).

[4] Equations (1)–(3) illustrate that in a simplified form, the partitioning of turbulent energy fluxes is a function of atmospheric demand (indicated by R_i , which is sensitive to R_n and δ_e), turbulent transport (R_a), “surface” resistance to water vapor transport (R_c), and air temperature (through s). Atmospheric demand is influenced by large-scale differences in climate, such that more continental sites will be drier and have a higher δ_e relative to R_n (large R_i), and maritime sites will be cooler and more humid, resulting in a lower δ_e relative to R_n (small R_i) [*Jarvis et al.*, 1976]. Atmospheric demand is also impacted by the development of the local atmospheric boundary layer, which may be closely coupled with surface fluxes. For example, sites with a high β may develop a warm and dry atmospheric boundary layer (higher R_i), which can result in a negative feedback on surface flux partitioning [*Jacobs and de Bruin*, 1992; *Raupach*, 1998; *Wilson et al.*, 1999]. Similarly, to the extent that the value of R_c mimics the integrated stomatal resistance of individual leaves [*Raupach and Finnigan*, 1988; *Paw U and Meyers*, 1989; *Kelliher et al.*, 1995], R_c may respond to short and long-term changes in atmospheric demand and soil moisture. For example, stomatal resistance often increases following an increase in δ_e [*Grantz*, 1990]. As a consequence of these feedbacks, sites with high atmospheric demand (δ_e and possibly R_i) might be expected have high values of R_c and vice versa. The extent to which these conceptual supply (R_c) and demand (R_i) terms in equation (3) compensate each other is an important factor influencing β .

[5] In this study we quantify the partitioning of LE and H during the warm season (mid-June to late August) at 27 FLUXNET sites over 66 site years. The goals of the study are (1) to evaluate the magnitude and partitioning of LE and H among various ecosystems, (2) to evaluate other bulk parameters, especially those derived from the big leaf equation, and (3) to use these bulk parameters to suggest the factors controlling β for different ecosystems and climates.

2. Materials and Methods

[6] More detailed descriptions of the FLUXNET methodology and summary studies from the network are given by *Aubinet et al.* [2000], *Baldocchi et al.* [2001a], and *Falge et al.* [2001]. The eddy covariance method is used to assess energy and trace gas fluxes between the biosphere and atmosphere at each site, using instrumentation typically placed several meters (shorter canopies) to a little more than 10 m (taller forests) above the canopy. Vertical flux densities of CO_2 (F_{CO_2}), latent (LE) and sensible heat (H) between vegetation and the atmosphere are proportional to the mean covariance between fluctuations of vertical velocity (w') and the respective scalar (c') (e.g., CO_2 , water vapor, and temperature) [*Paw U et al.*, 2000]. Measurements of above-canopy temperature, humidity and net radiation are also typically provided at FLUXNET sites.

[7] Data sets contributed by individual investigators to the FLUXNET database at Oak Ridge National Laboratory’s Data Archive Center (<http://daac.ornl.gov/FLUXNET>) contained eddy covariance and supporting environmental and meteorological data. The data from each of the contributing sites is separated by year. Each of these “site year” data sets contain mean half-hourly flux and meteorological data at eddy covariance stations across Europe and North America (Table 1). For each site, Table 1 shows the years of

Table 1. FLUXNET Sites Investigated in This Study

Site	State/Country	Period	LAI	Code	Coordinates	Species
<i>Deciduous Forests</i>						
Harvard	Massachusetts/USA	1992–1999	5.5	A	42°32'N 72° 11'W	Oak-Maple
WalkerBranch	Tennessee/USA	1995–2000	6	B	35°58'N 84° 17'W	Oak-Hickory
Hesse	France	1996–1999	6	C	48°40'N 7°50'E	European Beech
Vielsalm	Belgium	1996–1998	4.5	D	50°18'N 6°0'E	European Beech
Gunnarsholt	Iceland	1996–1998	1.4	E	63°50'N 20°13'W	Poplar
<i>Coniferous Forests</i>						
Tharandt	Germany	1996–1999	6	F	50°58'N 13°40'E	Norway Spruce
Norunda	Sweden	1996–1998	5	G	60°05'N 17°28'E	Norway Spruce, Scots Pine
Flakaliden	Sweden	1996–1998	2	H	64° 07' N 19°27'E	Norway Spruce
WeidenBrunnen	Germany	1996–1998	5	I	50°09'N 11°52'E	Norway Spruce
Hyytiälä	Finland	1996–1998	3	J	61°51'N 24°17'E	Scots Pine
Howland	Maine/USA	1996–1997	5.5	K	45°15' N 68°45' W	Spruce-Hemlock
Duke Forest	North Carolina/USA	1998 – 1999	5.2	L	35°52'N 79°59'W	Loblolly Pine
Bordeaux	France	1996 – 1997	2.8	M	44°0'N 0°5'E	Maritime Pine
North Boreas	Manitoba/Canada	1994 – 1997	4.5	N	55°54'N 98°30'W	Black Spruce
Aberfeldy	United Kingdom	1997 – 1998	8	O	56° 37' N 3°48' W	Sitka Spruce
Niwot Ridge	Colorado/USA	1999	4.2	P	40°02'N 105°33'W	Engelman spruce, Lodgepole pine, Subalpine Fir
<i>Mediterranean Climates</i>						
Blodgett Forest	California/USA	1997 – 1999	4.5	Q	38°53'N 120°37'W	Ponderosa Pine
Metolius	Oregon/USA	1996 – 1997	1.5	R	44°30'N 121°37'W	Ponderosa Pine
Castelporziano	Italy	1997 – 1998	3.5	S	41° 45'N 12°22'W	Quercus. ilex
Sky Oaks ^a	California/USA	1997 – 1998	1.1	T	33°22' N 116°37'W	Adenostoma-Ceanothus-Chaparral
<i>Crops</i>						
Bondville ^b	Illinois/USA	1997–1999	6.5	U	40°0'N 88°18'W	Corn/soybean alternate
<i>Tundra</i>						
Happy Valley	Alaska/USA	1994 – 1995	1.5	V	69°08'N 148°50'W	Eriophorum, Betula, Ledum, Vaccinium, Carex, Salix
Atkasuk	Alaska/USA	1999	1.5	W	70°29'N 157°25'W	Eriophorum, Sedges
Barrow	Alaska/USA	1998 – 1999	1.5	X	70°18'N 156°38'W	Carex, Dupontia, Arctophila
<i>Grasslands</i>						
LittleWashita	Oklahoma/USA	1998 – 1999	2	Y	34°58'N 97°59'W	Rangeland
Shidler	Oklahoma/USA	1997	2.8	Z	36°51'N 96°41'W	Tallgrass Prairie
Fort Peck	Montana/USA	1999	2	&	48°9'N 105°6'W	grassland

^aThe Sky Oaks site includes two subsites; site ear T1 is an older chaparral in 1997. T2 and T3 are younger chaparral in 1997 (T2) and 1998 (T3).

^bThe Bondville site alternates between corn and soybeans; U1 and U3 are years when corn was grown and U2 was a year when soybeans were grown. LAI is the maximum leaf area index for the site.

data analyzed in this study, along with a one-letter code used to denote the individual sites. The one-letter code and a subsequent numeral were a shorthand method for denoting the site and the year of measurement in this paper. For example, the code “A1” applies to site “A” (Harvard Forest) and the “1” denotes the first year of data for this site (1992 in this case). The sites in Table 1 are classified into six main groups, deciduous forests, coniferous forests, agriculture, tundra, grasslands and sites in Mediterranean climates. This last group is not a vegetation type, but it will be shown that the characteristics of Mediterranean climates, defined by relatively wet winters and dry summers, results in energy partitioning characteristics that differ from similar vegetation types in different climates.

[8] All sites are in the Northern Hemisphere, and the study period in this research was isolated to the late spring/summer period, between day 165 (mid-June) and day 235 (late August). Although seasonal and annual studies are also of unique interest [Baldocchi et al., 2001b], the summer season was chosen to constrain the study and to isolate the period when potential solar radiation and temperature, the driving forces for energy, carbon and water fluxes, are

usually greatest. The study period was also limited only to fluxes during the daytime, which was defined as the period when photosynthetically active radiation (PAR) exceeded zero. This strategy was chosen because the processes controlling surface exchange are often different during nocturnal periods [Mahrt, 1999] and because heat fluxes are typically much larger during the daytime than during the night. Furthermore, energy balance closure statistics and micrometeorological and chamber studies have indicated that the eddy covariance method is significantly less reliable during nocturnal periods [Lavigne et al., 1997; Law et al., 2001; Aubinet et al., 2000; Wilson et al., 2001]. Dew formation (negative LE) is not measured well because the sensors are often wet. If nocturnal data had been included in this study, β would likely have been slightly smaller than indicated because H is typically negative and LE is typically near zero at night.

[9] The cumulative heat fluxes (LE and H) and β were evaluated for each site year directly from mean diurnal trends, excluding nocturnal periods ($PAR = 0$). To estimate mean diurnal trends, flux and meteorological data from each site year between days 165 to 235 were converted into 48

half-hourly averages (i.e., up to 69 data points, one for each day, make up each of the 48 half-hourly averages). Data from nocturnal periods were subsequently excluded from the analysis. If data were available for more than 35% of all the daytime half-hour periods, the site year was kept for further analysis; otherwise, it was discarded because of inadequate sampling. Missing data were not empirically gap-filled using meteorological data, which has been useful in other applications [Falge *et al.*, 2001], to avoid assuming *a priori* relationships between environmental variables and flux estimates. Because data were not gap-filled using meteorological data in this study, errors or biases in the flux estimates were still possible because the averaging method in this study ignores periods of missing data. Although this bias was recognized and may be important in some individual cases, it was not expected to significantly alter the general results.

[10] In addition to β , the coefficient of *Priestly and Taylor* [1972] (α) was also computed. α is the ratio of seasonally cumulative daytime LE to seasonally cumulative daytime LE_{eq} (the equilibrium evaporation or $\epsilon R_n/(\epsilon + 1)$, where $\epsilon = s/\gamma$ [Jarvis and McNaughton, 1986]). α is often used to normalize evaporation to a climatological expectation [Baldocchi and Meyers, 1998]. When α is equal to 1.0, evaporation is at the thermodynamic equilibrium rate, which is the steady-state evaporation rate in a completely closed system [McNaughton and Jarvis, 1983; Wilson and Baldocchi, 2000]. The equilibrium rate is dependent only on the net radiation and temperature. Lower and higher values indicate evaporation rates that are lower and higher than the equilibrium rate, respectively.

[11] During midday periods (from 1000 to 1430 local standard time), the Penman-Monteith approximation to the big leaf equations was inverted to solve for the canopy resistance [Jarvis *et al.*, 1976; Shuttleworth *et al.*, 1984]:

$$R_c = (\epsilon\beta - 1)R_a + (\beta + 1)R_i \quad (5)$$

[12] R_a was computed as the sum of the resistance to momentum transport (u/u_*^2 , where u is the mean wind speed and u_* is the friction velocity) and an “excess resistance” for scalar fluxes, approximated by $4.9/u_*$ [Verma *et al.*, 1986]. The excess resistance is dependent on the structure of canopy elements [Garratt and Hicks, 1973], an effect that was not considered. However, the value of R_c was not highly sensitive to R_a , as found previously by Jarvis *et al.* [1976]. In the present study, hypothetical errors in R_a of 100% resulted in mean errors of about 10% in the computed value of R_c .

[13] Other climatological parameters computed were the mean midday (also from 1000 to 1430 local standard time) temperature (T_a), vapor pressure deficit (δ_e) and net radiation (R_n). A representative value of R_i was computed (equation (4)) using these midday estimates of δ_e and R_n . Because it was not reported or measured at many sites and to maintain quantitative consistency between sites, G was neglected in the computations of R_i at all sites. However, on a daytime basis G was usually less than 10% of R_n . Therefore the underestimation of R_i was not substantial at most sites, but the error resulting from this omission was probably more significant in sparse canopies such as the tundra locations where G was largest.

[14] One important quality control of eddy covariance data is energy balance closure. The sum of H and LE was generally less than the available energy ($R_n - G$) at FLUXNET sites, and there was a mean imbalance of about 20% in energy balance closure [Wilson *et al.*, 2001]. The mechanisms for the lack of closure are not fully understood, but a portion of this imbalance was likely an underestimation of heat storage (in the soil, air and biomass) components (G). Underestimates of these components would not affect the results in this study. There was no overall indication of differences in closure with β [Wilson *et al.*, 2001], which may suggest that H and LE were both systematically underestimated by similar relative fractions. Therefore it was possible that the relative partitioning between H and LE was generally accurate, but that the total magnitude of turbulent fluxes may be biased low.

3. Results and Discussion

3.1. Magnitude of Surface Fluxes and β

[15] For each site year, Table 2 shows the mean midday air temperature (T_a), vapor pressure deficit (δ_e), total daytime net radiation (R_n), and mean midday R_c , R_a , and R_i over the summer period. Also shown are seasonal estimates of β , α , and F_{CO_2} (the uptake of CO_2 , with positive values indicating uptake by the vegetation). F_{CO_2} in Table 2 is the mean value from the half-hour period when the mean F_{CO_2} was largest. The mean daytime sensible (H) and latent (LE) heat fluxes between days 165 and 235 are displayed graphically for each site year in Figure 1. Figure 1 also shows lines of constant β (dashed lines) and lines of constant total turbulent energy fluxes (sum of H and LE , solid diagonal lines). The mean daytime β during the summer period ranged from approximately 0.25 to more than 3.0 across a diverse sampling of ecosystems and climates. There are notable differences within vegetation types and between different years at the same site, many of which are dependent on unique site characteristics. However, broad vegetation and climate boundaries were subjectively placed on groups of data in Figure 1 based primarily on β , and to a lesser extent the magnitudes of H and LE . The β values across ecosystems were reasonably consistent with the very broad ranges summarized by Eugster *et al.* [2000]. In the following sections, surface energy partitioning is discussed for each of the broad vegetation and climate groups.

3.2. Deciduous Forests

[16] β was lowest at the agricultural site (Bondville, IL, USA, site U) and the deciduous forest sites (sites A–E), with a typical range between 0.25 to 0.5. At deciduous forests, β ranged from 0.11 at site year E1 (poplar forest in Iceland) to 0.73 at site year D1 (Beech forest in Belgium), a site that also contains some conifers (Douglas fir). Of the two deciduous forest sites with the longest records, both of which are in North America, β was lower (mean = 0.33) at the southern-most deciduous forest (Walker Branch, site B) relative to the forest 6.5° farther north (mean = 0.46) (Harvard Forest, site A).

3.3. Agriculture

[17] The low β and R_c in agricultural species was consistent with other studies [Davies and Allen, 1973; McGinn and King, 1990; Baldocchi, 1994], which probably results

Table 2. Summary of Climate and Big Leaf Parameters During the Period Between Day 165 and Day 235 for Each Site Year^a

Site	T_a	δ_e	R_n	R_a	R_c	R_i	β	α	F_{CO2}
<i>Deciduous Forests</i>									
Harvard (A1)	22.4	854	672	15	59	43	0.49	1.01	15.8
A2	22.7	1077	1005	12	70	36	0.38	0.69	18.2
A3	23.1	754	779	15	56	31	0.43	0.74	19.7
A4	23.3	951	1002	15	65	35	0.43	0.72	20.0
A5	20.9	832	867	13	63	33	0.46	0.77	19.1
A6	21.0	968	911	14	70	38	0.38	0.79	18.7
A7	22.2	838	956	14	64	31	0.56	0.69	16.3
A8	23.3	1149	1041	12	80	38	0.55	0.67	18.0
Walker Branch (B1)	27.6	1465	1050	22	101	50	0.43	0.63	17.1
B2	26.8	1188	1042	24	75	40	0.28	0.70	24.7
B3	27.1	1200	1092	24	74	39	0.31	0.71	20.8
B4	28.0	1312	1031	23	93	44	0.42	0.65	24.2
B5	27.4	959	991	26	67	33	0.33	0.74	22.3
B6	26.1	789	984	25	47	27	0.24	0.78	21.5
Hesse (C1)			878				0.34		15.8
C2			858				0.50		13.6
Viesalm (D1)			1017				0.73	0.53	14.4
Gunnarsholt (E1)	12.6	393	451	30	68	32	0.11	0.73	12.9
E2	13.6	426	498	33	96	34	0.54	0.61	8.2
Mean	23.0	947	901	20	72	37	0.42	0.72	18.0
<i>Coniferous Forests</i>									
Tharandt (F1)	19.3	995	918	13	82	40	0.69	0.70	18.2
F2	17.4	803	884	11	81	33	0.75	0.65	15.8
F3	19.0	1028	917	12	96	41	0.76	0.70	18.2
Norunda (G1)	18.0	973	862	13	94	45	0.79	0.67	20.5
G2	21.4	1333	963	13	132	53	0.90	0.57	16.3
G3	16.9	743	711	12	74	39	0.54	0.73	15.8
Flakaliden (H1)	17.3	891	812	19	105	42	0.79	0.68	8.0
H2	18.6	1027	906	23	151	45	1.07	0.48	6.2
H3	15.0	563	699	18	70	31	0.90	0.55	7.4
WeidenBrunnen (I1)	13.4	619	1034	15	109	21	2.02	0.36	
I2	16.6	759	804	16	130	33	1.12	0.40	9.6
I3	17.1	836	854	15	97	32	0.87	0.43	9.8
Hyytiälä (J1)	16.2	689	837	17	88	33	0.64	0.52	11.2
Howland (K1)			1058				0.97		12.5
K2			787				0.84		13.0
Duke (L1)	29.1	1874	1067	17	178	67	0.65	0.47	16.9
L2	30.9	2076	1102	18	142	66	0.52	0.58	15.0
Bordeaux (M1)	25.3	1787	1049	16	157	63	1.18	0.50	8.5
M2	25.3	1315	1029	17	81	49	0.46	0.73	13.0
North Boreas (N1)	19.3	1319	941	17	140	53	1.55	0.48	5.5
N2	18.0	1103	788	16	140	50	1.30	0.56	5.2
N3	18.6	1220	980	15	137	48	1.65	0.43	6.0
N4	18.0	1122	1006	15	148	43	1.77	0.45	6.1
Aberfeldy (O1)	16.0	652	697	21	190	36	2.20	0.38	14.8
O2	14.7	625	654	18	177	36	1.92	0.37	14.6
Niwot Ridge (P1)	14.8	910	1007	17	81	33	0.86	0.74	9.4
Mean	19.0	1053	899	16	120	43	1.07	0.55	11.9
<i>Mediterranean Climate</i>									
Blodgett (Q1)	23.1	1870	1144	26	116	60	0.47	0.79	11.5
Q2			1200				0.43		11.5
Q3	22.4	1814	1393	27	179	47	0.92	0.58	9.1
Metolius (R1)	24.0	2385	1264	16	271	65	1.51	0.37	8.6
R2	21.4	1806	1319	15	235	49	1.7	0.34	5.5
Castelporziano (S1)	23.0	957	1247	15	222	28	2.25	0.31	9.7
S2			1291				1.73		11.2
Sky Oaks (T1)	30.9	3862	1283	18	900	102	5.20	0.25	2.9
T2	29.5	3505	1301	21	2995	93	164.2	0.05	1.0
T3	28.9	3212	1326	27	317	85	0.86	0.47	7.3
Mean	25.4	2426	1277	21	654	66	17.9	0.40	7.8
<i>Crops</i>									
Bondville (U1)	26.2	1200	958	37	82	45	0.40	0.73	31.0
U2	27.1	1046	951	47	44	38	0.25	0.85	12.5
U3	26.5	1064	1002	28	55	37	0.28	0.82	41.6
Mean	26.6	1103	970	37	60	40	0.31	0.80	28.4

[20] The mean climatological resistance (R_i) was not significantly different between deciduous (37 s m^{-1}) and coniferous sites (43 s m^{-1}). Therefore the greater β at coniferous relative to deciduous sites was largely because R_c was lower in deciduous forests, and not because of systematic differences in R_i (see equation (4)). Similar mean values of R_i between groups do not suggest that there are no differences in climate between the coniferous and deciduous sites. However, the larger R_c at coniferous forests does suggest important ecological and physiological adaptations to the local environment, manifested by total leaf area and minimum stomatal resistance, parameters that typically have some correlation with R_c (z 1995). Generally smaller stomatal resistance (Körner *et al.*, 1978; Jones, 1992) and greater photosynthetic capacity [Wullschlegel 1993] and lower β [Barr *et al.*, 1997; Baldocchi *et al.*, 2000] have been observed in deciduous compared to coniferous forests. Within the conifer group there was considerable variation in δ_e , R_n , R_i , R_c , and some of these differences will be discussed later in relation to drought, climate and advection.

3.6. Mediterranean Sites

[21] The sites with Mediterranean climates were characterized by the largest values of R_n and $LE + H$ and a large variability in β (Figure 1 and Table 2). These sites generally had the greatest R_c (mean was 320 s m^{-1} even after excluding the particularly high value at site T2) and R_i (66 s m^{-1}). Sites in Mediterranean climates were most affected by the seasonal period chosen in this study, because vegetation in Mediterranean climates typically has greater productivity and evaporation rates in late winter/spring and/or autumn and not during summer [Valentini *et al.*, 1995; Radoglou, 1996; Anthoni *et al.*, 1999]. Mediterranean climates are typically defined by clear skies and high solar irradiance during this period of the year, but also by drying soil, which likely explains the large available energy and high R_c .

[22] Despite the generally high R_c at these sites, two of the years at the ponderosa pine plantation in Blodgett, CA, USA (site Q) had the highest total LE in the study and low β , although R_c was not particularly low (mean = 148 s m^{-1}); however, other sites within Mediterranean climates, the drier Ponderosa Pine forest at Metolius, Oregon, USA (site R), the broadleaved evergreen forest in Italy (site S) and the chaparral site at Sky Oaks, CA, USA (site T), have much lower LE and some of the highest values of β in the study despite the typically large values of R_i (i.e., R_c was very large and more than compensated for R_i in equation (3)).

3.7. Tundra Sites

[23] The mean flux partitioning characteristics of the tundra sites overlapped some of the coniferous sites in Figure 1, but β was generally slightly greater and LE and total turbulent energy fluxes slightly lower at tundra sites. Despite the greater β the mean R_c was lower (t test, $P < 0.01$, 27 degrees of freedom) at the tundra sites (71 s m^{-1}) than coniferous sites (120 s m^{-1}).

[24] The mean β of approximately one at tundra sites has been noted previously [Vourlitis and Oechel, 1997; Oechel and Vourlitis, 1998], but considerable site-to-site variability has also been observed, and McFadden *et al.* [1998] and Eugster *et al.* [2000] have quantified important differences in flux partitioning among vegetation types in arctic tundra.

One of the tundra sites (Barrow, AK, USA, site X) had a particularly low R_c (mean 45 s m^{-1}), but this site also had the lowest climatological resistance in the study (9 s m^{-1}), indicating that δ_e was low relative to R_n . As a consequence, β was actually the largest among the tundra sites. The Barrow site had a very shallow thaw zone and remained wet, resulting in a low R_c but also low δ_e (and R_i) (S. Brooks, personal communication). Advection from maritime sources can also be important in increasing β for some sites near northern coastal zones [Eugster *et al.*, 2000]. Because the vegetation is sparse at tundra sites, the energy balance is often influenced as much or more by local topography and hydrology, as well as the dynamics of mosses and the thaw zone, as by the overlying vegetation [Eugster *et al.*, 2000; Eaton *et al.*, 2001; Beringer *et al.*, 2001].

3.8. Synthesis Across Sites

[25] Previous discussion has focused on general differences in “big leaf” parameters and energy partitioning for groups of sites in this study. This section focuses on an overall evaluation of surface and climate controls of energy partitioning and the possible mechanisms for variability between sites. An assortment of complex biogeochemical, climatological, ecological and physiological processes control this variability, but general differences in partitioning can be inferred from the bulk parameters in the big leaf equation (equations (1)–(3)). Two important “big leaf” parameters determining β were R_c and R_i , which are conceptual indicators of surface supply and atmospheric demand (equation (3)). Figure 2 shows the values of R_c and R_i for all site years, with the exception of site years T1, T2 and Y1 (these were off-scale, but values are shown in Table 2). Figure 2 shows the approximate contours for β , which illustrates that variability in β across sites is the result of differences in both R_c and R_i (increasing R_i decreases β and increasing R_c increases β). The following discussion attempts to examine the important processes that resulted in differences in R_c and R_i between sites and what other sources of variation were important in controlling β .

[26] First, we address the sources of variability in R_c . In addition to the general trends in R_c discussed previously, such as the difference between coniferous and deciduous forests (conifer $R_c >$ deciduous R_c) and the difference between sites at Mediterranean climates and other sites (Mediterranean R_c usually greater), additional relationships may help explain differences between site years. R_c has been shown to be correlated with individual leaf stomatal resistance divided by the canopy leaf area index (LAI) [Kelliher *et al.*, 1995], although the relationship is expected to saturate at high LAI [Schulze *et al.*, 1995]. If tundra sites, which are often heavily influenced by the moss layer and dynamics of the thaw zone [Beringer *et al.*, 2001], are excluded, there was a weak negative correlation between R_c and the seasonal maximum LAI (Figure 3). The weak correlation ($r^2 = 0.22$) suggests that other physiological (e.g., individual leaf stomatal resistance [Kelliher *et al.*, 1995]) and nonphysiological (soil evaporation, canopy structure and turbulence [Raupach and Finnigan, 1988; Paw U and Meyers, 1989]) factors also influence R_c .

[27] Assuming R_c primarily reflects changes in stomatal resistance and canopy leaf area, R_c should reflect these general differences across ecosystems and climates. This

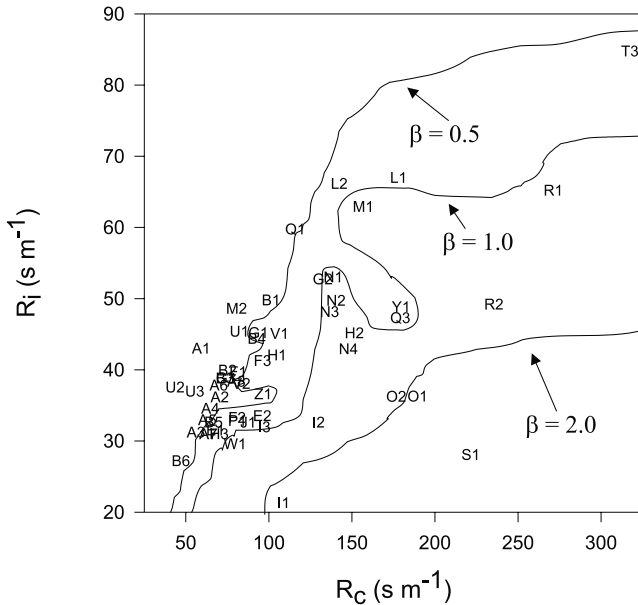


Figure 2. Relationship between surface resistance (R_c) and climatological resistance (R_i) for the site years in Table 1. The approximate isolines for β are shown. Site years T2, T3 and Y2 are off the scale of this graph.

includes the differences between vegetation types already discussed, but also differences within vegetation types. For example, The lower R_c and β at the Blodgett, CA ponderosa pine site (site Q) relative to the Metolius, OR ponderosa pine site (site R) was likely because Blodgett had twice the leaf area and three times the annual rainfall of the Oregon site (1630 versus 570 mm) [Goldstein *et al.*, 2000; Law *et al.*, 2001]. The greater leaf area at the Blodgett site is likely because of the greater annual rainfall, and more favorable autumn to spring climate, which helps offset the severe limitations imposed by summer drought. Therefore the value of R_i was fairly similar between these sites (sites Q and R in Figure 2) but R_c and β were not.

[28] As a second example, R_c and β at the boreal black spruce forest in Canada (site N) was greater than in the boreal forests of Europe (sites G, H, J). Low evaporation rates and high β have been noted previously in the boreal regions of Canada and Siberia [Fitzjarrald and Moore, 1994; Baldocchi and Vogel, 1997; Kelliher *et al.*, 1997; McCaughey *et al.*, 1997; Kelliher *et al.*, 1998; Baldocchi *et al.*, 2000], and may be related to low nutrient availability [Baldocchi and Vogel, 1997; Kelliher *et al.*, 1997] and photosynthetic capacity [Dang *et al.*, 1998]. Since needle area is not less at the Canadian site than at the European sites (Table 1), the result of the poor nutrient conditions at site N is apparently manifested by an especially high stomatal resistance per needle area [Dang *et al.*, 1998].

[29] R_c should also reflect interannual variations in physiological response, maximum leaf area and photosynthetic capacity at individual sites. Therefore one expected source of interannual variation in R_c and β at individual sites would be drought. Three case studies of drought have been published using the same data sets available in this research: a deciduous forest (site year B1 [Baldocchi, 1997]), a grassland (site year Y1 [Meyers, 2001]) and a ponderosa pine forest (site year Q3 [Goldstein *et al.*, 2000]). At all

three sites, this study indicates that R_c was at least 40% greater during these drought years (over 300% greater at grassland site). The considerable interannual variability of β and F_{CO_2} at the maritime pine stand in Bordeaux, France (site M) was primarily because of changes in R_c , which also may be the result of thin soils and quickly developing soil water deficits at that site [Berbigier *et al.*, 1996].

[30] A second important factor controlling β was the climatological resistance (R_i), which is a function of the magnitude of the atmospheric vapor pressure deficit relative to net radiation, or the “dryness” of the local environment relative to radiative forcing (equation (4)). In this study, there were differences between continental and maritime climates in the value of R_i (continental $R_i >$ maritime R_i). In continental regions, sites with a high R_c will likely warm and dry the atmospheric boundary layer and increase δ_e , R_i and the relative demand for evaporation [Raupach, 2000]. This is probably one reason why there was an overall positive correlation between R_c and R_i among the site years (Figure 2; $r^2 = 0.40$). At near coastal sites where advection from maritime sources is likely, such as the forests at Castelporziano, Italy (site S; approximately 15 km from Mediterranean Sea) and Aberfeldy, Scotland (site O; roughly 75 km from Atlantic), this feedback between R_c and R_i may be impeded because of advection of cooler, more humid air. At these two sites, R_i was low despite a high R_c (Figure 2). In addition, β at these two sites was greater than at other sites with similar values of R_c , probably because of the low R_i (Figure 2). The Duke forest, a continental site in the southeastern United States, is a stark contrast to these two sites. Duke forest (site L) had only a slightly lower R_c but a much lower β . This site had a high

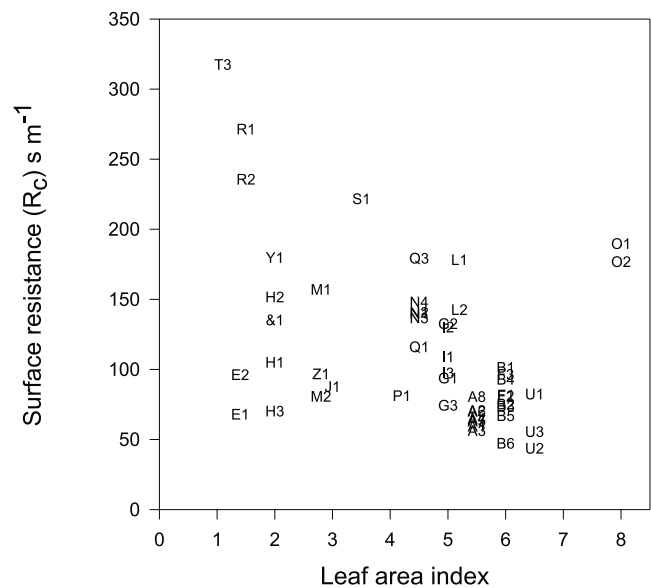


Figure 3. Relationship between leaf area index (LAI) and canopy resistance (R_c). LAI estimates were obtained from Law *et al.* [2001] and the FLUXNET web site. The letter-number codes refer to the sites in Table 1 (letter) and year of study at that site (number). Tundra sites were not included. Three site years (T1, T2, AND Y2) with very high R_c were off the scale on this graph. Enclosed circles denote subjective delineations between different vegetation types and climates, although there are exceptions.

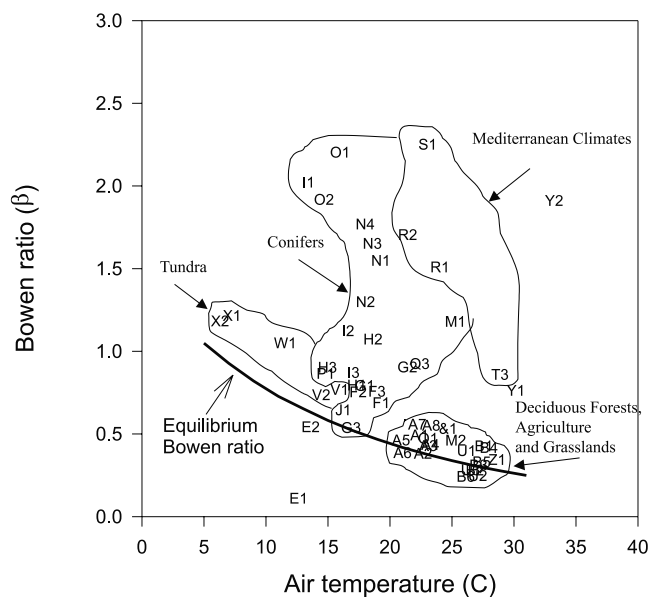


Figure 4. Relationship between mean daytime air temperature and Bowen ratio (β) for the site years in Table 1. Also shown is the equilibrium β . Site years T2 and T3 were off the scale of this graph. Enclosed circles denote subjective delineations between different vegetation types and climates, although there are exceptions.

climatic demand for evaporation (high δ_e and R_i). Similar reasoning explains why β was lower at the more continental Metolius, OR, USA site (R) than the more coastal sites O and S, despite a higher R_c .

[31] In addition to the feedback on the temperature and humidity of the atmospheric boundary layer, the positive correlation between R_c and R_i may also indicate the commonly observed stomatal response to high δ_e [Grantz, 1990]. These relationships may strengthen the positive feedbacks between R_c and R_i . For example, a low evaporation rate (high R_c and β) increases the atmospheric demand (δ_e and R_i), especially in continental regions, and a high demand (high δ_e and R_i) may prompt a diminished supply (higher R_c). Because R_c and R_i were not mathematically independent and are conceptualized bulk parameters, some correlation between the two may not be physical. Nevertheless, the relationship between climatic demand, surface resistance and β for different ecosystems helps illustrate the combined effects of large-scale climate, vegetation types and local feedbacks on energy partitioning.

[32] Because the contour lines for β are distorted, Figure 2 indicates that R_c and R_i do not account for all the differences in β between sites. Other factors, such as temperature, R_a , and the use of mean values of R_c , R_a and R_i instead of using instantaneous values in equation (3), are other sources of variability. The effect of temperature on the slope of the saturation vapor pressure deficit (s) and equilibrium evaporation (LE_{eq}) is one additional factor in determining β across climates that is often not considered. Assuming all other parameters in equation (3) are identical, β should decrease with temperature because s and LE_{eq} increase with temperature. Figure 4 shows equilibrium β , the value of β when evaporation is at thermodynamic equilibrium [Jarvis and McNaughton, 1986], as a function of air temperature.

As a result of the temperature sensitivity, the equilibrium β (value of β when $\alpha = 1$) decreases by a factor of three from 10°C to 30°C (from 0.80 to 0.27). Also shown in Figure 4 are the mean daytime air temperature and β at the FLUXNET sites. Although Figure 4 illustrates that temperature does not appear to be a major controlling factor in determining β across all sites, the effect of temperature on LE_{eq} may explain some of the more subtle differences between sites, such as between the southern and more northern deciduous forests in North America (sites A and B in Figure 4).

[33] The discussion on how resistance terms affect energy partitioning has focused primarily on the role of R_c and R_i rather than R_a , primarily because evaporation and energy partitioning are often much less sensitive to R_a (equation (3)) [Monteith, 1965]. Nevertheless, R_a , which showed some variability and was much greater in shorter vegetation, can affect the response of surface fluxes to R_c , δ_e and R_n [Jarvis and McNaughton, 1986] and the resulting β (equation (3)). For example, at the tundra, agricultural and grassland sites, the larger R_a should make β less sensitive to both R_c and R_i , and more sensitive to equilibrium evaporation. When evaporation is less than the equilibrium value ($\alpha < 1$), which was the typical scenario in this study (Table 2), LE increases (and β decreases) slightly with R_a because increasing R_a tends to push evaporation to the equilibrium rate [Monteith, 1965].

4. Conclusions

1. β was typically lowest at the agricultural and deciduous forest sites (0.25–0.50). In unstressed conditions, β was similarly low at grassland sites, but was likely sensitive to soil water content. There was variability in β at coniferous sites, but β was typically about twice the value compared to deciduous forests. The mean β of tundra sites was close to one. Sites in Mediterranean climates had the highest values of β in the study but also spanned a greater range and had greater available energy than the other groups.

2. The sources of variability in β between vegetation types, sites and years could be quantified using “big leaf” parameters. General differences between vegetation types were typically because of differences in the surface resistance to water vapor transport (R_c). However, differences between sites in the ratio of vapor pressure deficit to net radiation, manifested by the climatological resistance (R_i), also controlled β . Some coastal sites were influenced by their proximity to large bodies of water and advection, which reduced R_i and β . Other likely, though probably less important, factors controlling surface partitioning were temperature (through its effect on the slope of the saturation vapor pressure) and the atmospheric resistance to water and heat transport.

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- M. Aubinet, Unité de Physique, Faculté des Sciences Agronomiques de Gembloux, B-50 30 Gembloux, Belgium.
- D. D. Baldocchi and A. Goldstein, ESPM, University of California at Berkeley, Berkeley, CA 94720, USA.
- P. Berbigier, Unité de Bioclimatologie, INRA Bourdeaux, Gazinet, France.
- C. Bernhofer, Institute für Hydrologie und Meteorologie, 01737 Tharandt, Germany.
- H. Dolman, Alterra, P.O. Box 47, 6700 AC Wageningen, Netherlands.
- E. Falge and J. Tenhunen, Pflanzenökologie, University Bayreuth, 95440 Bayreuth, Germany.
- C. Field, Department of Plant Biology, Carnegie Institution of Washington, Stanford, CA 94305, USA.
- A. Granier, Unite Ecophysiologie Forestieres, Institut Nationale de la Recherche Agronomique, 54280 Champenoux, France.
- A. Grelle, Department of Ecology and Environmental Research, Swedish University of Agricultural Sciences, S-75007 Uppsala, Sweden.
- T. Halldor, Agricultural Research Institute, Department of Environmental Research, Reykjavik, Iceland.
- D. Hollinger, USDA Forest Service, Northeastern Forest Experiment Station, Durham, NH 03824, USA.
- G. Katul, School of the Environment, Duke University, Durham, NC 27708, USA.
- B. E. Law, College of Forestry, Oregon State University, Corvallis, OR 97331, USA.
- A. Lindroth, Swedish University of Agricultural Sciences, Department of Ecology and Environmental Research, SE-75007 Uppsala, Sweden.
- T. Meyers and K. B. Wilson, Atmospheric Turbulence and Diffusion Division, NOAA, P.O. Box 2456, Oak Ridge, TN 37831, USA. (wilson@atdd.noaa.gov)
- J. Moncrieff, Department of Ecology and Environmental Research, University of Edinburgh, Edinburgh, UK.
- R. Monson, Department of Environmental, Population, and Organismic Biology, University of Colorado, Boulder, CO 80309, USA.
- W. Oechel, Global Change Research Group, San Diego State University, San Diego, CA 92182, USA.
- R. Valentini, Department of Forest Science and Resources, University of Tuscia, I-01100 Viterbo, Italy.
- S. Verma, School of Natural Resource Sciences, University of Nebraska, Lincoln, Lincoln, NE 68583, USA.
- T. Vesala, Department of Physical Sciences, University of Helsinki, Helsinki, Finland FIN-00014.
- S. Wofsy, Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138, USA.